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ON THE ACCOUNTING OF SATELLITE-BORNE RADIATION MEASUREMENTS IN A NONADIABATIC MODEL OF ATMOSPHERIC MOTIONS

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ON THE ACCOUNTING OF SATELLITE-BORNE RADIATION MEASUREMENTS IN A NONADIABATIC MODEL OF ATMOSPHERIC MOTIONS * [Russian title]*

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Radiation heating or cooling, together with heat influx at the expense of turbulent heat conductivity and of hidden condensation heat play a part in the development of atmospheric flows which cannot be minimized. Computations by D. E. Martin [6] have shown that a further improvement of weather forecast methods require without doubt the accounting of nonadiabatic effects in the forecast schemes. It is however necessary to dispose to that effect of detailed informations concerning the space-time distribution of heat sources and outlets. Because of excessive complexity of this problem, it is desirable to study in the first place the effect of heat influx, averaged by the total atmosphere thickness, on the large-scale flows. Data obtained with the aid of meteorological satellites [5, 7] allow to find the value of this inflow directly.

Let us choose for a starting system of problem's equations the transport equation of velocity eddy, taking into account the turbulent friction forces

$$\frac{d_n(\Omega+l)}{dt} - \frac{g^2}{p_0^2}\frac{\partial}{\partial\zeta}\left(K\rho^2\frac{\partial\Omega}{\partial\zeta}\right) = l\frac{\partial\omega}{\partial\zeta} \qquad (1)$$

and the heat inflow equation

$$\frac{d_nT}{dt} - \frac{c^2}{R}\frac{\omega}{\zeta} = -\frac{g}{c_p p_0}\frac{\partial Q}{\partial\zeta} + \frac{\varepsilon}{c_p}. \qquad (2)$$

* K Uchetu rezul'tatov radiatsionnykh izmereniy so sputnikov v neadiabaticheskoy model atmosfernykh dvizheniy. [Presented by Academ. E. K. Fedorov on 11 April 1963].

[redacted]

Here Ω is the vertical component of the relative eddy velocity; l is the Coriolis parameter, g — the gravitation acceleration; K — the turbulent viscosity factor; ρ — the density of the air; T — the absolute temperature ; $c^2 = (\gamma_a - \gamma)R^2T/g$; γ and γ_a — the real and the dry-adiabatic vertical gradients of temperature respectively; c_p is the air's heat capacity at constant pressure; p is the pressure; p_0 — the standard pressure at sea level; $\zeta = p/p_0$.

The symbol d_n/dt denotes in both equations the individual derivative in time in a horizontal plane, and $\omega = d\zeta/dt$ is the analog of vertical velocity in the system of coordinates x, y, ζ, t.

The first addend of the right-hand part of equation (2) characterizes the inflow of heat to the unit mass of air at the expense of turbulent and radial heat exchange, σ/c_0 is the heat inflow at the expense of the concealed condensation heat. If we denote by D the turbulent heat flow, by B and A — respectively the heat fluxes directed upward and downward, by $F^{(1)}$ and $F^{(2)}$ — the ascending and descending fluxes of short-wave radiation, we have

$$Q = D + A - B - F^{(2)} - F^{(1)}. \qquad (3)$$

For boundary conditions along the vertical we shall admit that

$$\omega = 0 \quad \text{at} \quad \zeta = 0, \zeta = 1. \qquad (4)$$

In orderto pass in the given model from heat influxes to fluxes, we shall introduce for consideration a system of equations averaged by the vertical. Thus, integrating the equation (1) by from 0 to 1, and realizing the approximate accounting of the near-the ground friction according to I. A. Kibel' [1], we shall obtain in the assumption (4)

$$\overline{\frac{d_n(\Omega + l)}{dt}} + K^*\Omega_g = 0. \qquad (5)$$

Here $K^* \approx \frac{g\rho_0}{p_0}\sqrt{\frac{K_h l}{2}}$; K_h is the value of the turbulent viscosity factor at the upper boundary of the near-the ground sub-layer; Ω_g is the geostrophic eddy at sea level, $\overline{(\,)} = \int_0^1 (\,)\, d\zeta$.

The vertical velocity ω is determined from the equation (1):

$$\omega = \frac{1}{l}\left\{\int\limits_0^\zeta \frac{d_n(\Omega + l)}{dt} d\zeta - \frac{g^2}{p_0^2} K\rho^2 \frac{\partial\Omega}{\partial\zeta}\right\}. \qquad (6)$$

Substituting (6) into (2), averaging the obtained equation by the whole thickness of the atmosphere and adding then with the equation (6), we shall obtain

$$\frac{\overline{d_n T}}{dt} + \frac{c^2}{Rl}\left\{\overline{\varphi \frac{d_n(\Omega + l)}{dt}} - K^{\cdot}\Omega_g + \frac{g^2}{p_0^2}\overline{\frac{K_\rho^2}{\zeta}\frac{\partial\Omega}{\partial\zeta}}\right\} = \frac{g}{c_p p_0}(Q_0 - Q_1) + \frac{gLr}{c_p p_0}. \qquad (7)$$

Here $\varphi = 1 + \ln\zeta$. L is the concealed condensation heat, r is the quantity of precipitations.

As is shown by calculations, the quantity $\frac{g^2}{p_0^2}\overline{\frac{K_2^\rho}{\zeta}\frac{\partial\Omega}{\partial\zeta}}$ can be neglected by comparison with $K^{\cdot}\Omega_g$.

Although this is not compulsory for the given scheme, let us admit nevertheless the condition of atmosphere thermotropism [8]:

$$\nabla T(x, y, \zeta, t) = F(\zeta)\nabla\overline{T}(x, y, t), \qquad (8)$$

where ∇ is the gradient operator in the plane. The function $F(\zeta)$ may be determined by way of statistical processing of observation material. According to L. Berkovskiy [4] we have approximately $F(\zeta) = 2\zeta$. Utilizing the statics equation

$$T = -\frac{g\zeta}{R}\frac{\partial H}{\partial\zeta}, \qquad (9)$$

it is easy to obtain with the aid of (8) the correlations

$$(H)_{\zeta=1} = \overline{H} - \frac{R}{g}\overline{T}, \quad \nabla H = \nabla\overline{H} + \frac{R\Phi}{g}\nabla\overline{T}, \qquad (10)$$

where $\Phi = 1 - \int\limits_\zeta^1 \frac{F(\xi)}{\xi} d\xi$. Let us note that $\overline{F} = 1$ and $\overline{\Phi} = \overline{\varphi} = 0$.

If we consider the motion as quasigeostrophic, we shall obtain on the basis of the preceding $\overline{\frac{d_n T}{dt}} = \frac{\partial \overline{T}}{\partial t} + \frac{g}{l}(\overline{H}, \overline{T})$;

$$\overline{\frac{d_n(\Omega + l)}{dt}} = \frac{g}{l}\left\{\Delta \frac{\partial \overline{H}}{\partial t} + \left(\overline{H}, \frac{g}{l}\Delta\overline{H} + l\right) + \frac{R^2\overline{\Phi^2}}{g^2}\left(\overline{T}, \frac{g}{l}\Delta\overline{T}\right)\right\},$$

$$\varphi\overline{\frac{d_n(\Omega + l)}{dt}} = \frac{g}{l}\left\{\frac{R^2\overline{\varphi\Phi^2}}{g^2}\left(\overline{T}, \frac{g}{l}\Delta\overline{T}\right) - \frac{R\overline{\varphi\Phi}}{g}\left[\Delta\frac{\partial\overline{T}}{\partial t}\left(\overline{T}, \frac{g}{l}\Delta\overline{H}\right) + \right.\right.$$
$$\left.\left. + \left(\overline{H}, \frac{g}{l}\Delta\overline{T}\right) + (\overline{T}, l)\right\}.$$

The symbols ∇ and (a, b) designate, as usual, the Laplace operator on the plane and the jacobian (functional determinant) respectively.

Assembling the results obtained, we shall finally arrive at the folloxing system of model ecuations :

$$\Delta\frac{\partial\overline{H}}{\partial t} + \left(\overline{H}, \frac{g}{l}\Delta\overline{H} + l\right) + \frac{R^2\overline{\Phi^2}}{g^2}\left(\overline{T}, \frac{g}{l}\Delta\overline{T}\right) + K^*\left(\Delta\overline{H} - \frac{R}{g}\Delta\overline{T}\right) = 0; \qquad (11)$$

$$(\Delta - \mu^2)\frac{\partial\overline{T}}{\partial t} - \frac{R\overline{\varphi\Phi^2}}{g\overline{\varphi\Phi}} - \left(\overline{T}, \frac{g}{l}\Delta\overline{T}\right) + \left(\overline{T}, \frac{g}{l}\Delta\overline{H} + l\right) +$$
$$+ \left(\overline{H}, \frac{g}{l}\Delta\overline{T}\right) - \frac{g\mu^2}{l}(\overline{H}, \overline{T}) - \frac{gK^*}{R\overline{\varphi\Phi}}\left(\Delta\overline{H}\frac{R}{g}\Delta\overline{T}\right) =$$
$$= \frac{\mu^2 g}{c_p p_0}(Q_0 - Q_1) + \frac{\mu^2 g L r}{c_p p_0} \qquad (12)$$
$$\left(\mu^2 = \frac{l^2}{c^2\overline{\varphi\Phi}}\right).$$

The methods for the solution of this system of equations are well known (see [1]).

Let us pass now to the consideration of the right-hand part of the equation (12). The quantity Q_0 represents the radiation balance of the system " terrestrial surface — atmosphere", which may be computed directly by the data of satellite-borne measurements. The same quantity is computed with the view of forecasting from the well known formula

$$Q_0 = S_0(1 - \Gamma) - B_0, \qquad (13)$$

where S_0 is solar radiation flux entering the upper boundary of the atmosphere; Γ is the albedo of the system terrestrial surface- atmosphere; B_0 is the outgoing thermal radiation of the Earth and atmosphere.

In accordance with (3), we may obtain for Q_1 the expression

$$Q_1 = (1 - \alpha) S - E_{\text{эф}} - D_1. \tag{14}$$

Here α is the albedo of the underlying surface; S is the aggregate radiation flux; $E_{\text{эф}}$ is the effective radiation of the Earth's surface; D_1 is the turbulent heat flux near the ground.

The first two addends in the right-hand part of (14), just as the quantities B_0 and Γ in formula (13), are computed according to data of atmosphere vertical sounding, which, in principle, may be materialized with the aid of meteorological satellites. Entering in the computation of turbulent heat flux is the temperature of the underlying surface, for which average monthly temperatures of Ocean's surface, taken over numerous years, can be utilized for oceanic regions. To compute Q_1, meeting the conditions of dry land, it is appropriate to apply the equation of underlying surface's heat balance

$$-D_1 + (1 - \alpha) S - E_{\text{эф}} = \Pi - LE, \tag{15}$$

where Π is the heat flow in the ground, E — the rate of evaporation. Then, instead of (14), we shall obtain :

$$Q_1 = \Pi - LE. \tag{16}$$

If we utilize the data on radiation temperature of the ground surface T_π , obtained according to measurements from satellite, the heat flux to ground can be computed by the formula (2)

$$\Pi = \sqrt{\frac{c\rho\varkappa}{\pi}} \int_{-\infty}^{t} \frac{dT_\text{п}}{d\tau} \frac{d\tau}{\sqrt{t-\tau}}. \tag{17}$$

Here c, ρ and $\varkappa$ are respectively the specific heat capacity, the density and the soil's heat conductivity factor. For winter conditions and in regions where there is little evaporation, we shall have $Q_1 \approx \Pi$.

The flux Π may be found for the very same regions by using equation (15), provided the data on soil temperature are unavailable. As is well known, the turbulent heat flux is a function of wind velocity U and of the temperatures' ($T - T_{\Pi}$) difference [3]. In its turn, T_{Π} is determined by means of the flux Π as follows [2] :

$$T_{\text{n}} = \sqrt{\frac{1}{\pi \varkappa c \rho}} \int_{-\infty}^{t} \frac{\Pi \, d\tau}{\sqrt{t-\tau}}. \qquad (18)$$

Substituting (18) into (15), we shall obtain an integral equation for the determination of heat flux to soil according to the given radiation balance of the underlying surface.

$$D_1 \left(T - \sqrt{\frac{1}{\pi \varkappa c \rho}} \int_{-\infty}^{t} \frac{\Pi \, d\tau}{\sqrt{t-\tau}},\ U \right) + \Pi = (1 - \alpha) S - E_{\text{эф}}. \qquad (19)$$

Finally, for the computation of condensation heat fluxes we may utilize the factual data on the quantity of fallen precipitations.

The proposed scheme may be utilized as a forecast one, if we include in it the equation of moisture transport for the precalculation of fields of humidity, cloudiness and precipitations.

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in the name of A.I.Voyekov

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